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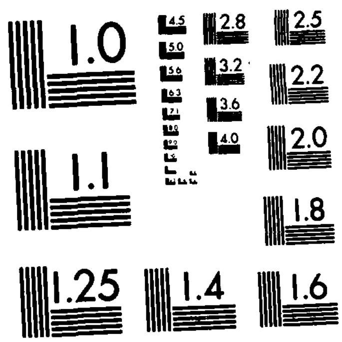
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Using the DWOPER routing model to simulate river flows with ice

Steven F. Daly and George D. Ashton

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The flow routing model of the National Weather Service entitled DWOPER (Dynamic Wave Operational Forecast Program) is examined with regard to the modifications required to include the effect of river ice on the flow variables of water level and discharge. Difficulties in modeling the ice effects are described. Example model output is presented showing the transient effects introduced by imposition or removal of the ice cover from an otherwise uncovered flow.

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PREFACE

This report was prepared by Steven F. Daly, Research Hydraulic Engineer, Ice Engineering Research Branch, Experimental Engineering Division, and Dr. George D. Ashton, Hydrologist, Snow and Ice Branch, Research Division, U.S. Army Cold Regions Research and Engineering Laboratory. The study was performed for the National Weather Service of the National Oceanic and Atmospheric Administration, Department of Commerce. Funding was provided through NOAA purchase order no. 01-8-M01-5105, Ice Jam Flood Predictions.

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CONVERSION FACTORS: U.S. CUSTOMARY TO METRIC (SI) UNITS OF MEASUREMENT

These conversion factors include all the significant digits given in the conversion tables in the ASTM Metric Practice Guide (E 380), which has been approved for use by the Department of Defense. Converted values should be rounded to the same precision as the original.

Multiply	By	To Obtain
British thermal unit	0.001055056	joule
foot	0.3048*	metre
$foot^2$	0.09290304*	$metre^2$
$foot^3$/second	0.02831685	$metre^3$/second
pound	0.4535924	kilogram
degrees Fahrenheit	$t_{°C} = (t_{°F} - 32)/1.8$	degrees Celsius

* Exact.

USING THE DWOPER ROUTING MODEL
TO SIMULATE RIVER FLOWS WITH ICE

Steven F. Daly and George D. Ashton

INTRODUCTION

The National Weather Service is responsible for forecasting water flows and water levels of the major rivers of the United States. Part of the forecast methodology involves calculation of the expected water levels and flows based on existing flows. To perform the calculations a variety of numerical models are used, all of which are based on the principles of open channel hydraulics. The most sophisticated of the currently used models is DWOPER (Dynamic Wave Operational Forecast Program) (Fread 1976). However, none of the models include the effects of ice on the hydraulics of the river flows and consequently on the water levels and discharges to be expected when ice is present in a river The effort reported here is a first attempt to include the effects of ice on the predicted water levels and flows in a model that includes unsteady (i.e. time-varying) effects.

DWOPER is an implicit dynam... .uting model that is being used by the National Weather Service to forecast floods and other flow conditions on major waterways. The model, which predicts flows and stages, is based on the one-dimensional differential equations of unsteady flow known as the Saint-Venant equations. These equations consist of an equation of continuity, which conserves the mass of the flow, and an equation of momentum, which conserves the flow momentum. Unlike earlier, simpler forecast models, DWOPER uses the complete unsteady flow equations which include the effects of frictional resistance, flow accelerations and surface slope. DWOPER also provides additional hydraulic information about flow along the waterways such as depth, flow cross-sectional area and top width, hydraulic radius, velocity, water surface slope, and energy slope.

DWOPER does not, however, include the effects of ice on the hydraulics of the flow. River ice can have a large effect on the stages and flows of a river. In fact, every ice process that occurs on a river produces a hydraulic transient of some magnitude. When an ice cover is present on a river, the frictional resistance of the cover to the flow must be taken into account. A certain portion of the channel cross section will also be blocked by the ice cover and be unavailable for flow. In the case of severe ice jams this portion may be quite large. The hydraulic radius, which is defined as the cross-sectional area of the flow divided by the wetted perimeter, must be modified to account for the increase in the wetted perimeter caused by the ice cover.

Also, the more "dynamic" ice effects should be taken into account. These include ice cover initiation, ice cover growth and evolution, ice jamming, ice cover breakup and ice cover movement. The hydraulic transients associated with these events can be very large. The conditions leading to the initial bridging of a river by floating ice are not well known, unless there is an obvious surface obstruction such as a floating ice boom or a dam. Factors that influence bridging include ice discharge rate, fragment size, open water width, ice concentration, lateral ice growth, air temperature and the ice transport capabilities of a given river reach.

On the other hand, the mechanics of the subsequent ice cover accumulation process have been studied to a much greater extent. Once an ice jam or bridging has begun, the ice cover accumulates through the arrival of ice fragments or blocks at the upstream end of the ice cover. The ability of the arriving blocks to resist entrainment in the water flowing under the existing ice can be estimated by knowledge of the thickness of the blocks and the velocity of the water. At higher velocities the blocks will underturn and be swept under the ice cover; they will deposit and thicken the cover somewhere downstream. The "mechanical stability" of the cover to withstand the forces applied it by the flowing water, gravity and wind can also be estimated. If the cover is not strong enough, it will collapse and thicken until it has enough internal strength to resist the applied forces.

Once an initial cover has been formed, further thickening results from loss of heat to the atmosphere. This thermal ice growth can be estimated

from air temperature and wind speeds by use of relatively simple formulations.

Ice cover breakup is a process that is not understood. Thermal and mechanical effects both play large, if only qualitatively understood, roles in it.

To summarize, river ice and the hydraulic features of river flow form an imperfectly understood feedback loop. The hydraulic characteristics of the flow affect the formation and evolution of the ice cover, which in turn affect the hydraulic characteristics of the flow. Ice effects on river flow can be quite dramatic, and serious efforts are being made to understand and quantify these effects and their consequences. It is the purpose of this paper to present a method of including the effects of river ice in DWOPER, to take some of the initial steps needed to include those effects, to show examples of the associated hydraulic transients, and finally to suggest additional steps that can be taken and indicate where further research is required.

CHANGES NEEDED TO ADAPT DWOPER TO ICE CONDITIONS

Hydraulic parameters

Two basic equations, the equation for conservation of the mass of the flow and the equation for conservation of the momentum of the flow, compose the one-dimensional differential equations of unsteady flow, known as the Saint-Venant equations, which are the theoretical bases of DWOPER. To fully account for the presence of an ice cover and the transport of the ice in the channel, both equations must be modified. The conservation of mass equation must be modified to not only conserve the mass of the water flow, but also the mass of the ice flow. The conservation of momentum equation must conserve the momentum and forces acting on the ice cover when it is moving, as well as the momentum and forces acting on the flow. However, at this stage of model development, we will assumed the ice cover to be either stationary or nonexistent. That is, the influence of the motion of the ice will not be included in either equation. Therefore, we will examine the conservation of momentum equation and only the modifications required to account for the presence of a stationary ice cover.

The law of conservation of momentum is used to derive the second Saint-Venant equation, i.e., the equation of motion or dynamic

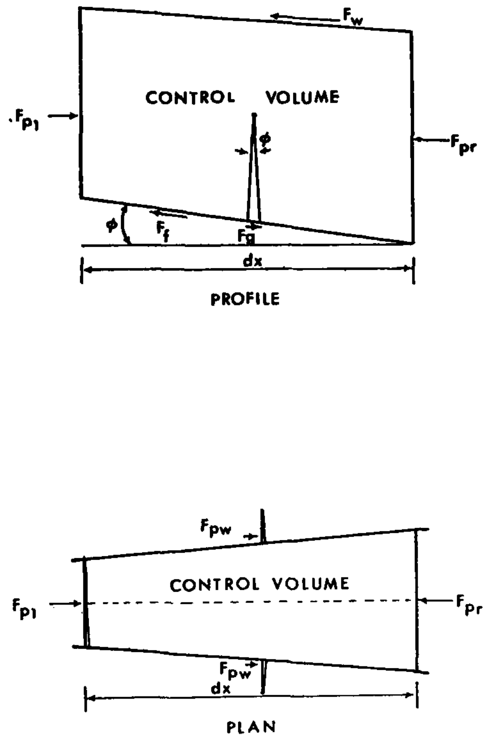

Figure 1. Definition sketch of control volume for derivation of momentum equation (from Fread 1976).

equilibrium. The conservation of motion is given by Newton's second law of motion, which may be stated as: The sum of the forces acting on the surface of the control volume plus the net rate of momentum entering the control volume equals the time rate of accumulation of momentum within the control volume.

The forces acting on the surface of the control volume of length dx shown in Figure 1 include 1) the gravity force due to the weight of the fluid F_g, 2) the force due to the frictional resistance along the channel bottom and sides F_f, 3) the force due to the shear stress produced by wind movement at the free surface of the control volume F_w, and 4) the unbalanced pressure force $F_{pl} - F_{pr}$; ϕ is the bottom slope angle and F_{pw} is the pressure force on the sides due to enlargement or constriction.

Of the four forces accounted for in the above description only the force due to the frictional resistance along the channel is directly affected by a stationary ice cover floating in the control volume. A stationary cover would also interfere with the wind force. However, the

wind force is usually either negligible or unknown and will not be included in this discussion.

The frictional resistance is manifested by a shear stress τ along the bottom and sides of the control volume. An empirical equation for open channel resistance, the Manning equation, is used to express the frictional resistance. The Manning equation (English units) is

$$V = \frac{1.486}{n_b} R^{2/3} S_f^{1/2} \qquad (1)$$

where V is the average velocity over the flow cross section, R is the hydraulic radius, S_f is the friction slope, and n_b is the Manning roughness coefficient of the channel bottom with dimensions of $ft^{1/6}$.

The conservation form of the momentum equation is

$$\frac{\partial Q}{\partial t} + \frac{\partial(\beta Q^2/A)}{\partial x} + gA\left(\frac{\partial y}{\partial x} + S_f - S_o\right) - \beta q\, V_x + \frac{\tau_w}{\rho} B = 0 \qquad (2)$$

where

Q = flow rate

t = time

β = momentum correction coefficient

x = distance along the channel

A = area of the flow cross section

g = gravity

y = depth of flow

S_o = bottom slope

q = lateral inflow rate with a velocity component V_x in the x direction

τ_w = wind shear stress at the water surface

B = width of flow

ρ = water density.

The friction slope S_f can be evaluated by the Manning equation. Substituting Q = VA into eq 1, we find

$$S_f = \frac{n^2 Q^2}{2.21 A^2 R^{4/3}} . \qquad (3)$$

The hydraulic radius, which is defined as A/P where P is the wetted perimeter, can be adequately approximated by the hydraulic depth $H = A/B$ for most natural river channels for which $B > 10y$. Therefore, we can write

$$S_f = \frac{n^2 Q^2}{2.21 A^2 H^{4/3}} \quad . \tag{4}$$

However, we must modify A, R and n in the above expression to account for the presence of the ice cover.

The cross-sectional area of the flow, A, must be reduced to account for the reduction of the cross-sectional area by the ice cover. If we assume a uniform ice cover of thickness η, the flow area A_i under the ice is

$$A_i = A - \frac{\rho_i}{\rho} \eta B \tag{5}$$

and A is determined as the cross-sectional area corresponding to the water level with ice present, ρ_i is the density of ice and B is the width of the cross section.

The hydraulic radius of a channel is defined as the cross-sectional area of the channel divided by wetted perimeter. The presence of an ice cover will increase the wetted perimeter by B, the width of the cross section. If we again assume that the depth is small when compared to the width, the hydraulic radius with an ice cover R_I is

$$R_I \simeq \frac{R}{2} \simeq \frac{H}{2} \quad . \tag{6}$$

The Manning roughness coefficient is an empirical measurement of channel roughness. It must be modified, relative to open water conditions, to account for the roughness of the underside of the ice cover. It should be emphasized that the roughness of the ice cover can vary both in space and through time. It can consist of both small scale (surface) drag and large scale drag when the underside is a jumbled pile of angular blocks and slabs. Many methods of determining the composite roughness, n_c, of an ice covered channel have been proposed. The following formula is easy to use and produces reasonable results as shown by Carey (1967) and Uzuner (1975):

$$n_c = \left(\frac{n_i^{2/3} + n_b^{2/3}}{2}\right)^{3/2} \tag{7}$$

where n_i is the roughness coefficient of the underside of the ice cover and n_b is the roughness coefficient of the channel bottom.

Equation 2, the momentum equation, can now be rewritten for the case of an ice-covered river, with the wind stress term dropped, as

$$\frac{\partial Q}{\partial t} + \frac{\partial(\beta Q^2/A_i)}{\partial x} + gA_i\left(\frac{\partial h}{\partial x} + S_f + S_o\right) - \beta q V_x = 0 \tag{8}$$

and eq 4 can be rewritten as

$$S_f = \frac{n_c^2 Q^2}{2.21 A_i^2 \left(\frac{H}{2}\right)^{4/3}} \ . \tag{9}$$

Ice parameters

Initiation of cover

As was noted in the introduction, unless there is an obvious surface obstruction such as a floating ice boom or a dam, the conditions leading to the initial bridging of a river by floating fragments are not well known. Factors that influence bridging include ice discharge rate, fragment size, open water width, ice concentration, lateral ice growth, air temperature, and the ice transport capabilities of a given river reach. Experience has shown that on given rivers the ice cover will initiate at about the same places every winter season. These sites are generally those places where the water surface slope changes from steeper to milder. The reasons for this are not entirely clear at this time; however, to successfully model a given phenomenon, clear and unambiguous criteria must be provided. Three approaches are proposed to achieve this: an arbitrary initiation, use of the Pariset and Hausser stability criteria, and a criterion based on surface concentration of floating ice.

Arbitrary initiation. A site where an ice cover will begin can be chosen arbitrarily. The choice can be based on experience, judgment or investigative need. It is a simple programming matter to select a location in the waterway where the ice cover will be initiated.

Pariset and Hausser stability criteria. In a now classic paper Pariset et al. (1966) developed criteria to determine the thickness of an ice cover formed by accumulation of floating ice fragments. Their analysis presumed that bridging had already taken place and examined the thickness of an ice cover that progresses upstream by accumulation. The essence of the analysis is a balance between the water forces on the floating ice cover and the resisting forces between the ice cover and the banks, with the thickness determined by the internal strength of the accumulation that is necessary to transmit the forces to the banks. While their criteria were developed for the case of ice cover progression rather than initiation, it may be argued that moving ice of the same thickness as the ice that they were describing will bridge across a channel and stop since that ice has the potential of resisting the forces. The criteria are examined in more detail later. For the moment, we simply note that the criteria relate the water depth, water discharge Q, channel width B and the hydraulic roughness C to a function of the ratio of the ice cover thickness η to depth of flow H in the form

$$X = \frac{Q^2}{BC^2H^4} = f(\eta/H) \ . \tag{10}$$

Pariset and Hausser used the Chezy coefficient C for the roughness parameter associated with the ice undersurface but it may easily be related to the Manning coefficient ($n = R^{1/6}/C$).

The function $f(\eta/H)$ is shown in Figure 2. If its value falls within or on the boundaries of the stable region, a cover is considered to be stable and initiation is presumed to have taken place. Experience gained during the present study showed this to be a very difficult criterion to satisfy; moving ice doesn't often fall within the stable region.

Concentration criteria. As ice first forms and moves down a river, it flocs together at the surface forming ice pans and hence is confined to the upper layer of the flow. If it remains only one layer thick, the ice must satisfy a continuity of surface flux of ice from one point to another relation. Expressing the surface concentration of ice as c, the surface velocity as V and the width as B, then at points 1,2 we find

$$c_1 \, V_1 \, B_1 = c_2 \, V_2 \, B_2 \ . \tag{11}$$

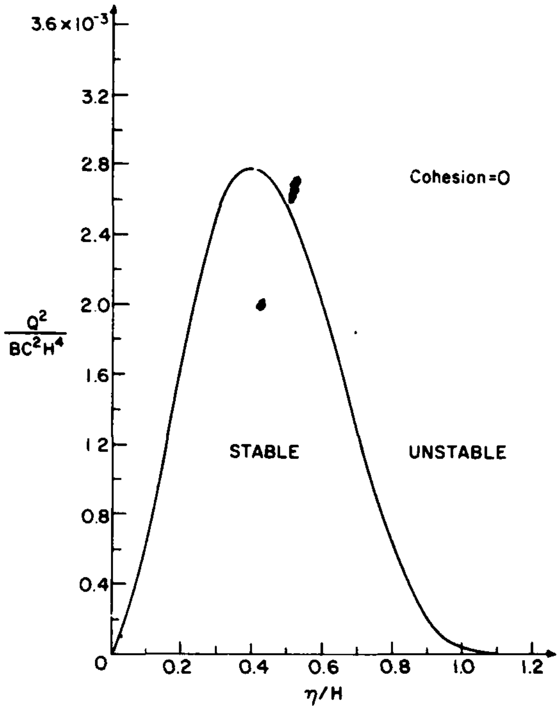

Figure 2. Stability diagram (after Pariset et al. 1966).

With knowledge of the variation in V and B along a reach, the concentration of ice of any point may be determined if the concentration is known at some upstream point. If this concentration is greater than the critical concentration at which bridging takes place, we then may presume that the ice cover has been initiated.

It is difficult, however, to determine the initial ice concentrations and the value of the critical concentration. Limited work has been done on the latter problem. Frankenstein and Assur (1972) speculated that broken ice will clog and come to a standstill when the surface concentration is about 0.7, but they offered no data to support this value. More recently Ackermann et al. (1979) analyzed the ice transport capacity of a river. They included the interaction of ice floes and were able to calculate the limiting capacity of a given channel to transport ice. In all cases examined the maximum ice discharge corresponded to concentrations of $0.6 < c < 0.7$ (Ackermann et al., in press). They also found by numerical analysis that the maximum ice discharge depended on the characteristic diameter of floes (larger floes clog more easily) and the slope (the milder the slope the less the required ice discharge). The most troublesome parameter for our purposes here is the floe diameter and this brings us full circle, i.e., even if criteria based on ice concentration (or floe diameter) are

available, it is currently beyond the capability of river ice mechanics to determine these parameters, other than by direct observation.

Ice cover progression

Once an ice cover has initially bridged a channel the cover may grow upstream. The stability of ice blocks arriving at the upstream end of the cover can be assessed using a simple equilibrium analysis. If the velocity of the approaching block is above some critical velocity V_c, the block will underturn and be swept under the cover; if below the critical velocity, the block will come to rest against the upstream end of the cover and the cover will progress upstream. Ashton (1974) found the critical velocity to be

$$\frac{V_c}{\left[g\eta_i\left(1-\frac{\rho_i}{\rho}\right)\right]^{1/2}} = \frac{2\left(1-\frac{\eta_i}{H}\right)}{\left[5-3\left(1-\frac{\eta_i}{H}\right)^2\right]^{1/2}} \tag{12}$$

where ρ_i and ρ are the densities of the ice and water respectively, and η_i is the thickness of the ice block. If the approach velocity is above the critical velocity for underturning, we can estimate the resultant thickness η of the jam by solving the expression determined by Pariset and Hausser (1961),

$$V = \left[2g\left(1-\frac{\rho_i}{\rho}\right)\eta\right]^{1/2} \left(1-\frac{\eta}{H}\right) . \tag{13}$$

Rates of ice cover growth can then be simply determined by balancing the incoming ice flux with the quantities required to increase the cover at this thickness.

Thickening

The ice cover may initially thicken further by mechanical collapse in response to the increasing forces of shear stress on the underside of the initial accumulation. Once this process is completed, the ice cover further thickens by thermal growth.

As the ice cover progresses upstream by accumulation at the thickness η, which is found by application of eq 13 using the local variables of velocity and depth, it may not have enough internal strength to transmit

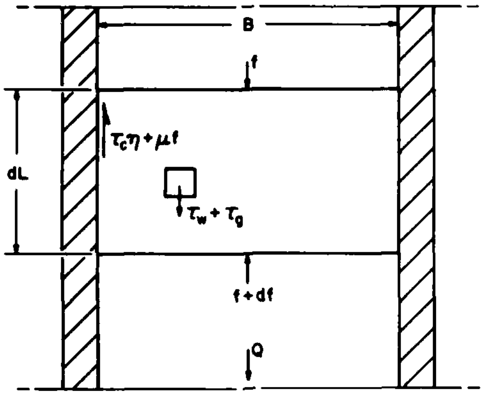

Figure 3. Definition sketch for analysis of ice accumulation stability.

the shear stresses acting on its underside to the banks. This is particularly true for so-called "wide" rivers. If the ice cover is not strong enough it will thicken further by a process often termed "shoving" until it has the competence to transmit the forces to the banks. Pariset and Hausser (1961) analyzed this situation as follows.

A differential length dL (Fig. 3) of an ice cover of thickness η and width B experiences an overall force balance of

$$Bdf + 2\left(\tau_c\eta + \mu f\right)dL = \left(\tau_w + \tau_g\right)BdL \tag{14}$$

where τ_c is the cohesion of the ice to the banks, μ is a type of ice friction coefficient on the banks, τ_w is the water shear stress on the underside of the cover, and τ_g is the gravity force per unit area acting on the cover along the slope. Equation 14 may be integrated to yield

$$f = \left[\frac{B(\tau_w + \tau_g)}{2\mu} - \frac{\tau_c\eta}{\mu}\right]\left[1 - \exp\left(-\frac{2\mu L}{B}\right)\right] \tag{15}$$

where the boundary condition that f = 0 at L = 0 has been used.

A distinction can be made in eq 15 between wide and narrow rivers. If the first bracketed term is negative the river is characterized as narrow because the ice cover can carry the shearing forces of the water and the gravitational forces without further thickening.

In eq 14 and 15, τ_w can be calculated from open channel hydraulics, as can τ_g. Pariset et al. (1966) found from field measurements that μ had a value of about 1.28 and the product $\tau_c\eta$ varied from 75 to 92 lb ft^{-1} for the St. Lawrence River (but it varied for other rivers).

If the river is wide the cover will thicken by shoving until the internal resistance is sufficient to carry the forces to the banks.

Once an ice cover has been initiated on a river and is not moving, it thickens further by thermal growth. Incremental thickening can be quite well estimated by stepped integration of

$$\frac{d\eta}{dt} = \frac{1}{\rho_i \lambda} \left[\frac{-(T_a - T_m)}{\frac{\eta}{k_i} + \frac{\eta_s}{k_s} + \frac{1}{h_{ia}}} \right] \tag{16}$$

where

η = ice thickness

t = time

ρ_i = density of ice

λ = heat of fusion

η_s = thickness of any snow layer on the surface

k_i and k_s = thermal conductivities of ice and snow

T_a = air temperature

T_m = 32°F, the melting point of ice

h_{ia} = heat transfer coefficient applied to the difference between the top surface temperature and the air temperature.

Equation 16 is based on a quasi-steady heat conduction analysis (Ashton 1980) that is more than adequate, considering the uncertainty in pragmatically available values of the input parameters. The value of h_{ia} ideally is calculated using detailed energy budget analyses; as a practical matter h_{ia} may be estimated and is of the order of 3 to 4 Btu ft^{-2} hr^{-1} °F^{-1}.

HYDRAULIC TRANSIENTS ASSOCIATED WITH ICE PROCESSES

Introduction

Hydraulic transients are associated with all ice processes that occur on a river. The magnitude of the associated transients can be very large or very small, depending on a variety of conditions. In this report we investigate the potential magnitude of the hydraulic transients by examining two processes: the freezeup and the breakup of a uniform ice cover. An idealized prismatic channel, with a variable slope and hydraulic roughness, is used to examine the hydraulic transients associated with these two processes. One of the difficulties of studying the unsteady effects of ice

development is the almost complete lack of reliable field data. Therefore, our first purpose is to determine what is important and what is not.

An important concept in the study of transients caused by an ice cover is the concept of normal or uniform depth. The depth of a uniform flow is the normal depth. A given river discharge will be at normal depth when the slope of the water surface, S_f, is equal to the slope of the channel bottom; therefore

$$S_f = S_o \tag{17}$$

where S_o is the slope of the channel bottom. For open water flow at uniform depth, we know from eq 3

$$Q = \frac{1.486}{n} AR^{2/3} S_o^{1/2} \tag{18}$$

and that for wide channels

$$Q = \frac{1.486}{n} (BH_N)(H_N)^{2/3} S_o^{1/2} \tag{19}$$

where H_N is the normal depth. Rearranging, we find

$$H_N = \left[\frac{Q\ n}{1.486\ B\ S_o^{1/2}}\right]^{3/5} . \tag{20}$$

Similarly, for an ice covered river, assuming that the hydraulic roughness coefficient of the ice undersurface is the same as the bottom and that we can ignore temporarily the portion of the channel blocked by ice, we find

$$Q = \frac{1.486}{n} (BH_N)\left(\frac{H_N}{2}\right)^{2/3} S_o^{1/2} \tag{21}$$

or

$$H_{NICE} = 1.32 H_N . \tag{22}$$

Thus, the presence of an ice cover increases the normal depth by approximately 32%. If the area of the channel blocked by the ice is included (i.e. the depth is called the free water surface), it has been found that an accurate description for the normal depth is

$$H_{NICE} = 1.32 H_N + \frac{\rho_i \eta}{\rho} \tag{23}$$

where $\rho_i \eta / \rho$ represents the displacement effect of the ice cover.

While it is seldom that a given discharge is actually flowing in a channel at its normal depth, it is the change of normal depth as a result of an ice cover forming or breaking up and the actual water depth relative to the resultant normal depth that determines the potential magnitude of the associated hydraulic transient. The ultimate magnitude of the hydraulic transient, however, will be determined by the downstream constraints, the slope of the river channel, the length of the channel, and the duration of the ice processes. It is hard to disentangle the relative importance of these factors; however, the channel slope is probably the most important factor. The steeper a channel, the shorter will be the distance over which the transition to a new equilibrium depth occurs in response to a change in the downstream conditions (an example is shown in Fig. 4). Hence, for a given downstream constraint, the slope of the channel largely determines the extent of influence of an ice formation or ice breakup event. The larger transients are expected on the steeper channels.

Breakup

Figure 5 illustrates the transient response of three prismatic rectangular channels with varying channel slopes to breakup of a uniform ice cover. The downstream elevation in each case is held constant at the ice covered normal depth, and the upstream discharge held constant. The ice

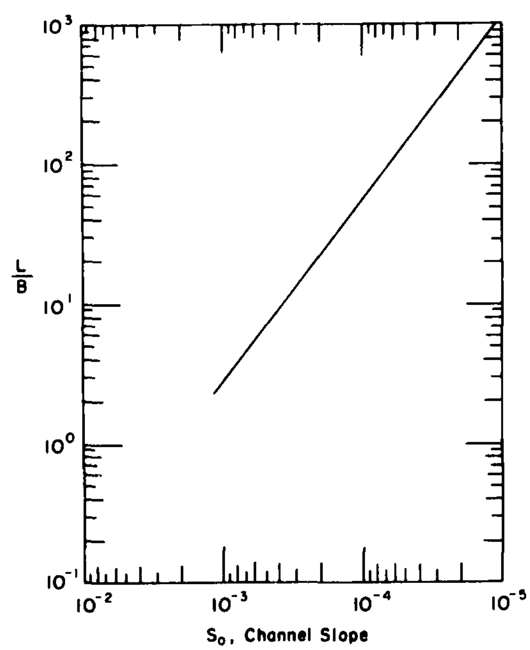

Figure 4. Example calculation of distance upstream at which 99% of normal depth with ice H_{NICE} is attained with downstream constraint fixed at open water normal depth H_N.

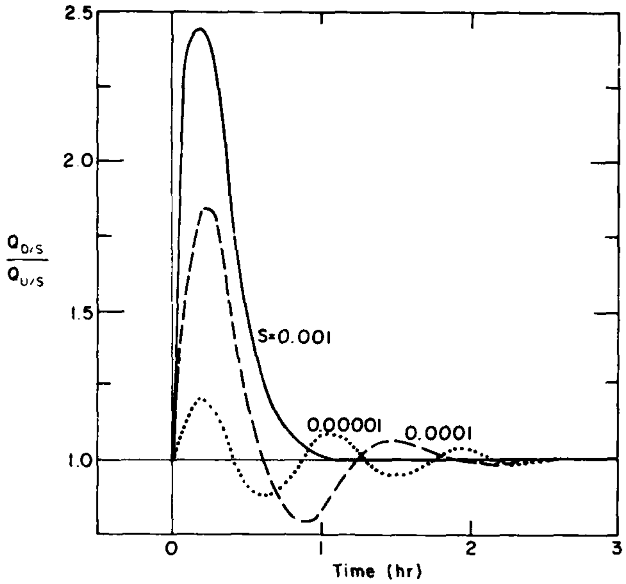

Figure 5. Transient response to ice breakup for prismatic channels of three different slopes.

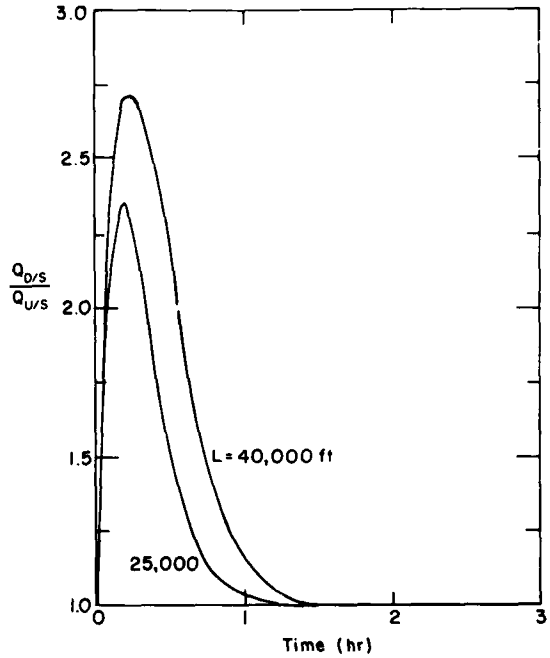

Figure 6. Effect of length of channel on transient response to breakup (S = 0.001).

cover was assumed to instantaneously cease being a hydraulic influence. (In the numerical simulation this was done by instantaneously changing the hydraulic radius to the depth of flow beneath the ice cover, reverting from the composite roughness to the more usual open channel roughness, and adjusting the cross-sectional area.) Only for the steeper channels does this approximation approach reality. The magnitude of the transient response does indeed increase with channel slope. The milder channels also tend to show more reflections of the transient, and the oscillations of the discharge can be noted for these cases. The effect of varying the length of the channel on the transient response can be seen in Figure 6. The larger transient is due to the cumulatively longer release of storage water caused by the change in resistance. In general, the longer and steeper the channel is, the greater the transient response to ice breakup.

Freezeup

We can reverse the above process to study the transient response to the freezeup of a river channel. Figure 7 illustrates the transient response of the same three prismatic rectangular channels to the ice cover freezeup. The downstream elevation in each case is held constant at the

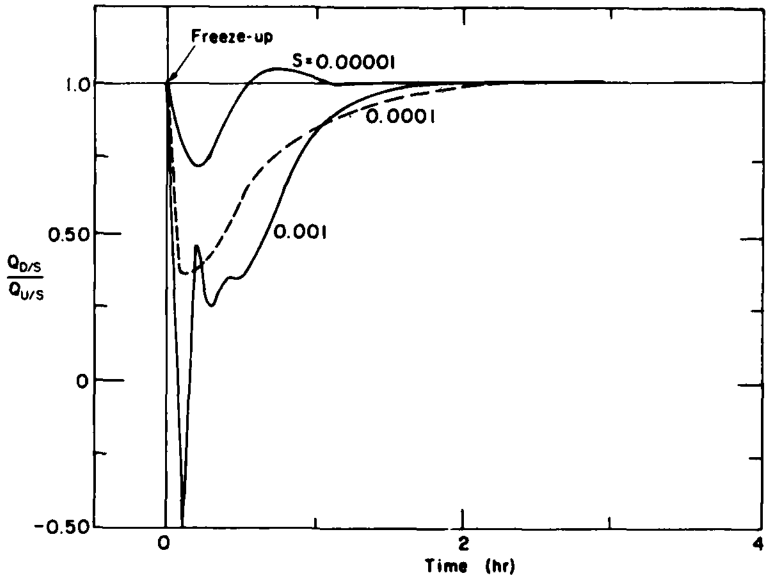

Figure 7. Transient response to instantaneous freezeup of rectangular prismatic channels of three different slopes ($Q = 300{,}000\ ft^3\ s^{-1}$, constant downstream elevation at open water normal depth).

open water normal depth, and the upstream discharge held constant. The hydraulic influence of the ice cover was assumed to appear instantaneously over the entire length of the channel. We know that river freezeup is generally a gradual process that happens over a length of time. Thus, these examples are extreme cases to examine the potential magnitude of the hydraulic transients that may take place. Immediately apparent is the numerical instability shown for the steepest river. This will be more fully discussed in the next section. Again, the magnitude of the transient response increases with channel slope. Many different processes can combine to cause a river to freezeup. The river velocity is known to be a critical parameter, and the lateral growth rates, ice inflows from upstream, and meteorological conditions all contribute to the overall process. The situation is greatly simplified in this simulation, but again, it allows examination of potential magnitudes.

MODIFICATIONS TO DWOPER PROGRAM

The modifications to DWOPER are very straightforward. Basically, the terms of the conservation of momentum equation are modified as described earlier. The hydraulic influence of the ice cover is "felt" by the channel through the modification of three parameters: the composite hydraulic

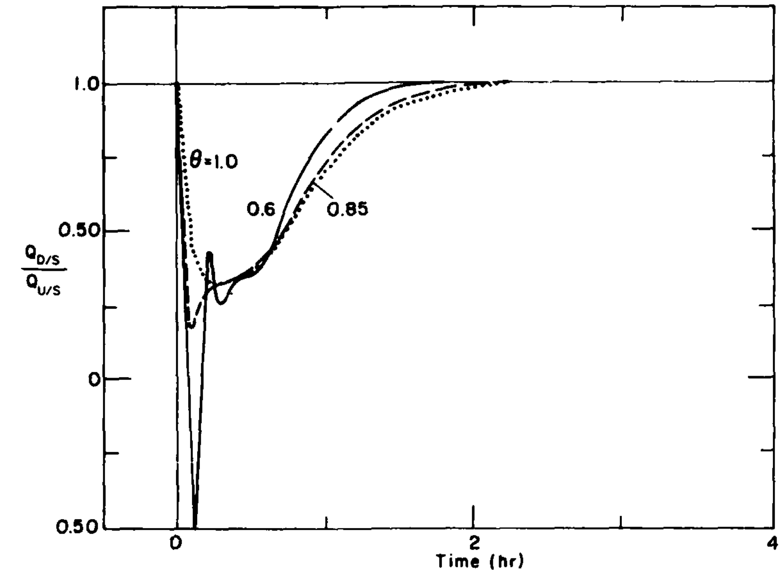

Figure 8. Effects of weighting factor on stability for a case of instantaneous freezeup (S = 0.001).

roughness of channel, the cross-sectional area and the hydraulic radius. Whenever the presence of an ice cover was indicated by an ice thickness greater than zero, these parameters were changed to reflect that presence.

Several runs were made to test the effect of varying the weighting factor, θ. This parameter determines the positioning of spatial derivatives and nonderivative terms between adjacent time lines. Generally, a value for θ of 0.55 to 0.6 is used, and this was the value used in the previous examples. However, an instability problem was noted for the steepest channel during freezeup. No instability problems were noted for the case of ice breakup. Problems of instability of the numerical solutions are a result of the nonlinearity of the various terms of the governing equations. Frictional resistance is a very nonlinear term that assumes more importance for the steeper channels. The presence of an ice cover increases frictional resistance. Therefore, the freezeup on a steeper channel would be a most likely process to cause instability problems to appear. Figure 8 shows the results of several runs that were made with varying values of θ. We can see that the instabilities did not entirely disappear until θ was set equal to a value of 1.00. However, the front of the wave was slightly lagged and the magnitude damped. This would suggest that a variable θ that can change in anticipation of or in response to the changing ice conditions may be appropriate.

SUMMARY

This report is a step toward an eventual goal of accurately modeling the processes of river ice and the associated transient responses. We have shown that by modifying the equation of the conservation of the momentum of the flow by modifying three parameters -- the hydraulic roughness, cross-sectional area and the hydraulic radius -- we can indeed produce hydraulic transients that qualitatively, at least, resemble the real world. We are a long way, however, from starting our simulation in the fall, and predicting the ice processes and resultant transients through to the spring breakup period. The following points summarize our findings.

1. The transient response of channels to the process of ice freezeup and ice breakup can be demonstrated by modifying the appropriate parameters in the equation of the conservation of momentum of the flow. In general, the steeper the channel, the greater the transient response.

2. A more fully developed and robust simulation of the transient response could probably be achieved by including the conservation of ice discharge and the conservation of ice momentum and forces. A variable θ, the weighting factor, that could be varied in response to the changing ice conditions would be appropriate.

3. To accurately simulate the ice cover growth, the physics and mechanics of lateral ice growth, longitudinal ice floe growth, ice cover initiation by bridging or arching, and ice cover breakup must be more fully understood.

4. The analytical investigation of the transient response of channels and rivers to ice processes must be broadened and deepened. Very little work has been done in this area.

5. The collection of field data must be improved because very little field data are available. While the situation is improving, the lack of methods for remote sensing of ice thicknesses is severely hampering the field effort.

6. The effect of ice jams, as opposed to uniform ice cover, should be investigated. Hydraulically, a jam can be defined as any ice cover that produces a steep gradient relative to the channel slope across the length of the cover. The breakup of jams approaches the dam break problem as a worst case. The rapid formation of jams produces a response that closely resembles that produced by partial closure of a sluice gate as a worst

case; positive waves propagating in the upstream direction and negative waves propagating downstream from a sudden jam site have been measured in the field.

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